THE IERS SPECIAL BUREAU FOR THE OCEANS

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The oceans have a major impact on global geophysical processes of the Earth. Non-tidal changes in oceanic currents and ocean-bottom pressure have been shown to be a major source of polar motion excitation and also measurably change the length of the day. The changing mass distribution of the oceans causes the Earth's gravitational field to change and causes the center-of-mass of the oceans to change which in turn causes the center-of-mass of the solid Earth to change. The changing mass distribution of the oceans also changes the load on the oceanic crust, thereby affecting both the vertical and horizontal position of observing stations located near the oceans. Recognizing the important role that non-tidal oceanic processes play in Earth rotation dynamics and terrestrial reference frame definition, the International Earth Rotation Service has recently created a Special Bureau for the Oceans in order to facilitate research into these and other solid Earth geophysical processes affected by the oceans.

Introduction

The Earth's rotation, encompassing both the rate of rotation as well as the terrestrial location of the rotation pole, is not constant but changes on all observable time scales from subdaily to secular. This rich spectrum of observed Earth rotation changes reflects the rich variety of astronomical and geophysical phenomena that are causing the Earth's rotation to change, including, but not limited to, ocean and solid body tides, atmospheric wind and pressure changes, oceanic current and bottom pressure changes, torques acting at the core-mantle boundary, and post-glacial rebound.

Recognizing the rich variety of processes affecting the Earth's rotation, the International Earth Rotation Service (IERS) has recently created a Center for Global Geophysical Fluids in order to help relate dynamical properties of the atmosphere, oceans, and core to motions of the Earth, including its rotation (Dehant et al., 1997). As part of the Global Geophysical Fluids Center, the IERS Special Bureau for the Oceans (SBO) is responsible for collecting, calculating, analyzing, archiving, and distributing data relating to non-tidal changes in oceanic processes affecting the Earth's rotation, deformation, gravitational field, and geocenter. The oceanic products available through the IERS SBO are produced primarily by general circulation models of the oceans that are operated by participating modeling groups and include oceanic angular momentum, center-of-mass, bottom pressure, and torques.

Core SBO Products

Oceanic Angular Momentum

In the absence of external torques, the angular momentum of the entire Earth does not change. However, if the angular momentum of one component of the Earth, such as the oceans, changes, then the angular momentum of the other components of the Earth must change in order for the angular momentum of the entire Earth to remain constant.

Since observing stations are located on the Earth's crust, observations of the Earth's rotation determine the rotation of the solid Earth. As the angular momentum of the solid Earth is exchanged with that of the other components of the Earth, the rotation of the solid Earth will change. Interpreting the observed changes in the rotation of the solid Earth, which encompasses changes in both the length-of-day and in the location of the Earth's rotation axis with respect to the crust (polar motion), therefore requires knowledge of the changes of the angular momentum of the other components of the Earth, such as the oceans. Further discussion of various aspects of oceanic angular momentum and Earth rotation changes can be found in the studies of Brosche and Sündermann (1985), Brosche et al. (1990, 1997), Frische and Sündermann (1990), Ponte (1990, 1997a), Ponte and Gutzler (1991), Eubanks (1993), Dickey et al. (1993), Ponte and Rosen (1994), Bryan (1997), Segschneider and Sündermann (1997), Dickman (1998), Furuya and Hamano (1998), Marcus et al. (1998), Ponte et al. (1998), and Celaya et al. (1999).

By definition, the time-dependent angular momentum vector $\mathbf{L}(t)$ of the oceans is given by:

$$\mathbf{L}(t) = \mathbf{L}_{\mathbf{mass}}(t) + \mathbf{L}_{\mathbf{motion}}(t) = \int_{V_o(t)} \rho(\mathbf{r},t)\, \mathbf{r} \times [\, \boldsymbol{\omega} \times \mathbf{r} + \mathbf{u}(\mathbf{r},t)\,]\, dV \tag{1}$$

where the integral is taken over the entire volume $V_o(t)$ of the oceans (which, in general, is a function of time) and $\mathbf{r}$ is the position vector of some oceanic mass element of density $\rho(\mathbf{r},t)$ that is moving with Eulerian velocity $\mathbf{u}(\mathbf{r},t)$ with respect to a terrestrial reference frame that: (1) is fixed to the solid Earth, (2) has an origin located at the center of the Earth, and (3) is rotating with angular velocity $\boldsymbol{\omega}$ with respect to an inertial reference frame. The total angular momentum of the oceans is the sum of two parts: (1) a mass term $\mathbf{L}_{\mathbf{mass}}(t)$ which will change as the mass distribution of the oceans changes, and (2) a motion term $\mathbf{L}_{\mathbf{motion}}(t)$ which will change as the direction and strength of the oceanic currents change. In general, $\boldsymbol{\omega}$ is not constant in time but exhibits variations in both magnitude (related to changes in the length of the day) and direction (related to polar motion). However, these changes are small, and for the purpose of deriving expressions for the angular momentum of the oceans, it can be assumed that the Earth is uniformly rotating at the rate Ω about the z-coordinate axis: $\boldsymbol{\omega} = \Omega\, \hat{\mathbf{k}}$. In this case, the Cartesian components of the mass term of the angular momentum of the oceans can be written as:

$$L_{x,mass}(t) = -\Omega \int_{V_o(t)} \rho(\mathbf{r},t)\, r^2 \sin\phi \cos\phi \cos\lambda \, dV \tag{2}$$

$$L_{y,mass}(t) = -\Omega \int_{V_o(t)} \rho(\mathbf{r},t)\, r^2 \sin\phi \cos\phi \sin\lambda \, dV \tag{3}$$

$$L_{z,mass}(t) = \Omega \int_{V_o(t)} \rho(\mathbf{r},t)\, r^2 \cos^2\phi \, dV \tag{4}$$

where ϕ is North latitude and λ is East longitude. Similarly, the Cartesian components of the motion term of the angular momentum of the oceans can be written as:

$$L_{x,motion}(t) = \int_{V_o(t)} \rho(\mathbf{r},t)\, [r \sin\lambda \, v(\mathbf{r},t) - r \sin\phi \cos\lambda \, u(\mathbf{r},t)] \, dV \tag{5}$$

$$L_{y,motion}(t) = \int_{V_o(t)} \rho(\mathbf{r},t)\, [-r \cos\lambda \, v(\mathbf{r},t) - r \sin\phi \sin\lambda \, u(\mathbf{r},t)] \, dV \tag{6}$$

$$L_{z,motion}(t) = \int_{V_o(t)} \rho(\mathbf{r},t)\, r \cos\phi \, u(\mathbf{r},t) \, dV \tag{7}$$

where $u(\mathbf{r},t)$ is the eastward component of the velocity and $v(\mathbf{r},t)$ is the northward component.

Oceanic Center-of-Mass

In the absence of external forces, the location of the center-of-mass of the entire Earth does not change. However, if the center-of-mass of one component of the Earth, such as the oceans, changes, then the center-of-mass of the other components of the Earth must change in order for the center-of-mass of the entire Earth to remain constant.

Artificial satellites of the Earth, which orbit about the center-of-mass of the entire Earth, are tracked by a global network of stations located on the surface of the Earth's crust. From these tracking measurements, the offset of the entire Earth's center-of-mass from the center-of-figure of the network of tracking stations, known as the geocenter, can be deduced. Since the tracking stations are located on the Earth's solid surface, changes in the location of the geocenter can be taken to represent changes in the location of the center-of-mass of the solid Earth. Interpreting the observed offset of the entire Earth's center-of-mass from that of the solid Earth (or, strictly speaking, from the geocenter) requires knowledge of the changing locations of the center-of-mass of the other components of the Earth, such as the oceans. Fortunately, the center-of-mass of the oceans can be easily estimated from products of general circulation models of the oceans (Dong et al., 1997; Chen et al., 1999).

By definition, the time-dependent position vector $\mathbf{r}_{\mathrm{cm}}(t)$ of the center-of-mass of the oceans is given by:

$$\mathbf{r}_{\mathrm{cm}}(t) = \frac{1}{M(t)} \int_{V_o(t)} \rho(\mathbf{r},t)\, \mathbf{r} \, dV \tag{8}$$

where $\mathbf{r}$ is the position vector of an oceanic mass element of density $\rho(\mathbf{r},t)$, the integral extends over the entire volume $V_o(t)$ of the oceans, and $M(t)$ is the total mass of the oceans which is assumed here to be time-dependent and which can be computed by integrating the density field:

$$M(t) = \int_{V_o(t)} \rho(\mathbf{r},t)\, dV \tag{9}$$

The Cartesian coordinates x_{cm}, y_{cm}, and z_{cm} of the center-of-mass of the oceans can therefore be computed from its time-dependent density distribution by:

$$x_{cm}(t) = \frac{1}{M(t)} \int_{V_o(t)} \rho(\mathbf{r},t)\, r \cos\phi \cos\lambda \, dV \tag{10}$$

$$y_{cm}(t) = \frac{1}{M(t)} \int_{V_o(t)} \rho(\mathbf{r},t)\, r \cos\phi \sin\lambda \, dV \tag{11}$$

$$z_{cm}(t) = \frac{1}{M(t)} \int_{V_o(t)} \rho(\mathbf{r},t)\, r \sin\phi \, dV \tag{12}$$

Ocean-Bottom Pressure

The changing mass distribution of the oceans causes the load on the ocean floor to change. As the oceanic crust and mantle yields to this changing load, the position of stations located on the crust near the oceans will change, and the geoid will deform (vanDam et al., 1997). Computing station displacements and perturbations to the geoid caused by ocean loading requires knowledge of the pressure $p_b(\phi,\lambda,t)$ at the bottom of the oceans caused by the weight of the overlying oceanic mass. This can be estimated by vertically integrating the time-dependent density distribution $\rho(\mathbf{r},t)$ of the oceans:

$$p_b(\phi,\lambda,t) = g \int_{-H(\phi,\lambda)}^{\eta(\phi,\lambda,t)} \rho(\mathbf{r},t)\, dz + p_a(\phi,\lambda,t) = g\rho_o \eta(\phi,\lambda,t) + g \int_{-H(\phi,\lambda)}^{0} \rho(\mathbf{r},t)\, dz + p_a(\phi,\lambda,t) \tag{13}$$

where η is the instantaneous sea level height, H is the ocean depth, and g is the gravitational acceleration which is assumed here to be constant everywhere within the oceans. The final expression in Eq. (13) is obtained by assuming that the surface layer of height η has constant density ρ_o. For completeness, the surface atmospheric pressure $p_a(\phi,\lambda,t)$ has been included in the expression for ocean-bottom pressure. However, see the Discussion Section below for some comments on the response of the oceans to atmospheric pressure and the effect of this response on oceanic angular momentum, center-of-mass, and bottom pressure.

Extended SBO Products

Gravitational Field Coefficients

The changing mass distribution of the oceans causes the Earth's gravitational field to change, an effect which will soon be accurately measured by the CHAMP and GRACE satellite missions. Interpreting the observed gravitational field changes over the oceans requires knowledge of the ocean-bottom pressure (Wahr et al., 1998). This can be estimated from the density distribution of the oceans by Eq. (13). Expanding the time-dependent ocean-bottom pressure field in spherical harmonics (using appropriate spatial averaging functions, if desired) yields time-dependent gravitational field coefficients that can be compared to those soon-to-be observed by CHAMP and GRACE.

Oceanic Torques

Torques exerted by the oceans on the bounding solid Earth effect changes in the rotation of the solid Earth. These torques are due to either frictional stresses acting on the solid boundaries of the ocean basins, or due to pressure gradients acting on topographic features of the ocean basins. Investigating how the oceanic angular momentum is transferred to the solid Earth requires knowledge of these various torques (Ponte, 1990; Ponte and Gutzler, 1991; Ponte and Rosen, 1994; Bryan, 1997; Segschneider and Sündermann, 1997).

Auxiliary SBO Products

Oceanic Mass

General circulation models of the oceans that are formulated using the Boussinesq approximation conserve volume. Artificial mass variations in such models can be introduced if there are density changes due to internal mixing or imposed surface heat fluxes. Since volume is conserved, the changing density will artificially change the mass of the ocean model. Mass conservation can be imposed on Boussinesq ocean models by adding a uniform layer to the surface that has the appropriate time-dependent thickness (Greatbatch, 1994; Mellor and Ezer, 1995; Dukowicz, 1997; Ponte, 1999, Sec. 2). To compute the required thickness of this surface layer, the mass variation exhibited by the Boussinesq ocean model must be known. This can be calculated using Eq. (9).

Land-Ocean Mask and Ocean-Bottom Topography

In order to impose mass conservation upon Boussinesq ocean models, the surface area of the model ocean must be known as well as the mass variation. In addition, the bottom topography of the ocean model must be known in order to compute the ocean-bottom pressure (see Eq. 13) as well as the torques acting on the solid Earth caused by both friction and by pressure gradients acting on topographic features of the ocean bottom.

Discussion

Products of general circulation models of the oceans can be used to compute the oceanic angular momentum, center-of-mass, and bottom pressure as outlined above. If these models are formulated using the Boussinesq approximation, then as discussed above, mass conservation should be imposed by adding a uniform layer to the ocean-surface that has the appropriate time-dependent thickness. Of course, the effect of this mass-conserving layer upon the oceanic angular momentum, center-of-mass, and bottom pressure needs to be taken into account by using the corrected sea-surface height in Eqs. (1–13). However, if the oceanic general circulation models (OGCMs) are not formulated using the Boussinesq approximation, but are rather formulated to conserve mass, such as that currently under development at the NASA Goddard Institute for Space Studies (GISS), then no such correction for mass conservation needs to be made.

The response of the oceans to atmospheric pressure fluctuations must be considered when computing its effect on the Earth's rotation, deformation, gravitational field, and geocenter. The effect of this response on the oceanic angular momentum, center-of-mass, and bottom pressure will be automatically included if these quantities are computed from the products of an OGCM that is forced by atmospheric pressure as well as by wind stress and heat flux. However, if atmospheric pressure is not included in the forcing applied to the OGCM, then this effect must be computed separately. In general, the oceans can be expected to respond dynamically to atmospheric pressure forcing, in which case an ocean model must be used to compute the response. However, for atmospheric pressure changes occurring on time scales greater than a few days, the response of the oceans can be approximated by that of an inverted barometer (ib) wherein sea level $\eta_{ib}(\phi,\lambda,t)$ is related to surface atmospheric pressure $p_a(\phi,\lambda,t)$ by:

$$\eta_{ib}(\phi,\lambda,t) \;=\; \frac{-1}{g\rho_o}\,[p_a(\phi,\lambda,t) - \bar{p}_a(t)] \tag{14}$$

where $\bar{p}_a(t)$ is the time-dependent average pressure over the global oceans. The inverted barometer approximation is regularly used when computing the effect of atmospheric pressure fluctuations on the Earth's rotation (see the chapter on the IERS Special Bureau for the Atmosphere elsewhere in this Technical Note). Using the inverted barometer approximation, the equilibrium response of the oceans to surface pressure fluctuations can be taken into account in the expression for ocean-bottom pressure by incorporating Eq. (14) into (13):

$$p_b(\phi,\lambda,t) \;=\; g\rho_o\eta_d(\phi,\lambda,t) \;+\; g\int_{-H(\phi,\lambda)}^{0} \rho(\mathbf{r},t)\,dz \;+\; \bar{p}_a(t) \tag{15}$$

where $\eta_d(\phi,\lambda,t)$ is the sea level height caused by dynamic (non-equilibrium) ocean processes and is defined by $\eta_d(\phi,\lambda,t) = \eta(\phi,\lambda,t) - \eta_{ib}(\phi,\lambda,t)$. Further discussion of various aspects of the response of the oceans to atmospheric pressure fluctuations and the applicability of the inverted barometer approximation can be found in the studies of Dickman (1988, 1998), Ponte et al. (1991), Wunsch (1991), Ponte (1992, 1993, 1994, 1997b, 1999), Tai (1993), vanDam and Wahr (1993), Hoar and Wilson (1994), Fu and Pihos (1994), Woodworth et al. (1995), Gaspar and Ponte (1997), and Wunsch and Stammer (1997).

SBO Product Availability

A World Wide Web home page and anonymous ftp site for the IERS Special Bureau for the Oceans is currently under development. When they are operational, the data sets being archived by the IERS SBO will be available through them. Until then, the data sets can be obtained upon request from R. Gross by sending electronic mail to Richard.Gross@jpl.nasa.gov. In addition, a Fortran subroutine for computing the IERS SBO core products from the results of oceanic general circulation models is also available upon request from R. Gross.

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